Geological controls on slope deformations in the Köfels rockslide area (Tyrol, Austria)

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Abstract

The largest crystalline mass movement in the Alps, the Holocene Köfels rockslide (Tyrol, Austria), was re-investigated in order to study the influence of lithologies and geological structures on deep seated gravitational slope deformations. The slide debris detached from an east-facing slope and left a giant niche, where the geological and morphological conditions point to different types of failure mechanisms. According to the lithologies encountered in the scarp and accumulation area, the main source of the rapid Köfels rockslide is represented by a several hundred metres thick orthogneiss complex embedded in paragneisses of the Austroalpine Ötztal basement unit. Layering and foliation of both the orthogneisses and the surrounding paragneisses are orientated unfavourable to promote slope failures. Rather this slope collapse was clearly structurally predisposed by fault-related valley-deepening and the coalescence of different brittle fracture systems. The key-discontinuities that reduced here the rock mass strength and enabled the failure are two main fracture sets: in the orthogneisses, fracture set og#1 comprises the well-known 25-35° E-dipping sliding planes that daylight subparallel to the slope, set og#2 forms a cluster of 60-80° E- to NE-dipping fractures. In the paragneisses, the fractures measured show a similar but more scattering spatial distribution. Concerning the slope collapse, the newly compiled geological map indicates that the size of the southern scarp area, which is entirely made up by paragneiss series, obviously differs from the volume of according lithologies in the accumulation area. Based on these findings and the scarp morphology, considerable parts of the scarp niche, i.e. areas made up by incompetent paragneisses, were not involved in the rapid Köfels rockslide event but characterised by slowly creeping slope deformations. Furthermore the geological field surveys and remote sensing data, i.e. optical, laserscanner and InSAR images, show that the present-day head scarp is widely affected by post-failure slope deformations. As a major feature, en-echelon orientated steeply E-dipping fractures highly influence the rock mass anisotropy, reduce the strength of the orthogneisses and cause further slope instabilities. Also in the adjacent paragneisses, distinct double-crested ridges, several secondary scarps and extension zones are signs of ongoing rocksliding. As a whole, the Köfels site is an outstanding example of how deep-seated rock slope failures and their potential evolution are controlled by the lithological inventory and the structural setting.

Die größte Massenbewegung im Kristallin der Alpen, die Holozäne Felsgleitung von Köfels (Tirol, Österreich), wurde neuerlich untersucht um den Einfluss von Lithologien und geologischen Strukturen auf tiefgründige gravitative Hangdeformationen zu erfassen. Die Gleitmassen entstammen einem nach Osten gerichteten Hang und hinterließen eine riesige Ausbruchsnische, in der die geologischen und morphologischen Verhältnisse auf verschiedene Typen von Versagensmechanismen hinweisen. Entsprechend den im Abbruch- und Ablagerungsgebiet aufgeschlossenen Lithologien wird die Hauptquelle dieser raschen Felsgleitung von einem mehrere Hundert Meter mächtigen Orthogneis Komplex aufgebaut, umgeben von Paragneisen der Austroalpinen Ötztal Grundgebirgsdecke. Lagenbau und Schieferung von beiden Einheiten, den Orthogneisen und den umgebenden Paragneisen sind entsprechend ungünstig orientiert um Hangversagen zu fördern. Vielmehr war diese Massenbewegung klar strukturell angelegt durch störungsgebundene Talübertiefung und Vernetzung von verschiedenen spröden Bruchsystemen. Die entscheidenden Trennflächen, welche hier zur Reduktion der Gebirgsfestigkeit und zum Hangversagen führten, sind zwei Hauptkluftsets: in den Orthogneisen umfasst Kluftset og#1 die bekannten 25-35° E-fallenden und subparallel zum Hang ausstreichenden Gleitflächen, Set og #2 bildet einen Cluster von 60-80° E- bis NE-fallenden Störungs- und Kluftflächen. In den Paragneisen weisen die gemessenen Trennflächen eine ähnliche, jedoch mehr streuende räumliche Verteilung auf. Bezüglich des Hangkollapses zeigt die neu kompilierte geologische Karte, dass die Größe des südlichen Abbruchgebietes, welches vollständig von Paragneis-Serien aufgebaut wird, offenbar vom Volumen entsprechender Lithofaziestypen im Ablagerungsgebiet abweicht. Basierend auf diesen Erkenntnissen und der Morphologie des Abbruchgebietes waren bedeutende Abschnitte der Abbruchnische, i. e. die aus inkompetenten Paragneisen aufgebauten Bereiche, nicht an der raschen Felsgleitung von Köfels beteiligt, sondern von langsam kriechenden Hangdeformationen geprägt. Weiters zeigen die geologischen Geländeaufnahmen und Fernerkundungsdaten, i. e. optische, Laserscanner und InSAR Aufnahmen, dass die heutige Abrisskante von weit verbreiteten post-Abbruch-Hangdeformationen erfasst wird. Hauptmerkmale sind staffelartig angeordnete steil E-fallende Bruchzonen, die eine deutliche Gebirgsanisotropie bewirken, die Festigkeit der Orthogneise herabsetzen und zu weiteren Hanginstabilitäten führen. Auch in den angrenzenden Paragneisen weisen ausgeprägte Doppelgratbildungen, zahlreiche sekundäre Abbruchzonen und Dehnungsbereiche auf anhaltende Felsgleitungen hin. Insgesamt stellt das Ge-

biet von Köfels ein hervorragendes Beispiel dar, wie das tiefgründige Versagen von Felsflanken und deren mögliche Entwicklung von den lithologischen Verhältnissen und den strukturellen Rahmenbedingungen geprägt werden

1. Introduction

Some of the largest rock slope failures in the Alps occur spatially clustered in the area of the Upper Inn valley - Ötz valley (Tyrol, Austria). They feature various types of rockslides and rockfalls, with deposition volumes ranging between some 10-1.000 million m^3 and with run-out distances extending up to several kilometres (Abele, 1974; Prager et al., 2008). In order to better understand the geological causes, progressive failure mechanisms and slope deformation behaviours, several of these landslides in different settings have been investigated recently. The tasks performed cover geological and geophysical field surveys of the landslide deposits, radiometric age dating of events as well as detailed mapping of the scarp areas (e.g. Weißflog, 2007; Zangerl et al., 2007; Pagliarini, 2008; Prager et al., 2009). Compiled results indicate that deep-seated gravitational slope deformations can be attributed to site-specific lithological, structural and morphological predispositions as well as to complex and polyphase interactions of several geological, hydrogeological and time-dependent rock mechanical processes. Among these are different long-term rock strength degrading processes e.g. stress redistribution due to morphological changes and subcritical fracture propagation, as well as shorter termed variable triggering factors e.g. seismic activities and groundwater fluctuations. Furthermore the geological field survey results showed a strong relation between rock mass strength and slope deformation be-haviour. In the Tyrolean Central Alps (Austria), the majority of deep-seated slope instabilities are rockslides that are encountered in incompetent metamorphic rocks such as paragneisses, micaschists and phyllites. Whereas some of these slopes are stabilised to date or characterised by a very low to non-detectable activity, others are still creeping at low velocities of up to some centimetres per year (e.g. Tentschert, 1998; Leobacher and Liegler, 1998; Chwatal et al., 2005; Zangerl et al., 2007). In contrast, gravitational slope deformations in competent rocks such as orthogneisses and amphibolites are mainly characterised by episodic but high-velocity failure events comprising rockfalls, rapid rockslides and rock avalanches (cf. Hungr and Evans, 2004).

In view of these findings, the largest Alpine landslide in crystalline bedrock, the prominent Köfels event (Ötz valley, Tyrol) was recently re-analysed and investigated geologically. Due to the enormous volume of >3.2 km^3 (Brückl et al., 2001) and the occurrence of “pumice” (Pichler, 1863), these rockslide deposits have been subject to scientific research for about 150 years. Numerous studies were concerned with sedimentological and morphological features (Ampferer, 1939; Klebelsberg, 1951; Heuberger, 1966; Hermanns et al., 2006), the genesis of fused rocks (Preuss, 1974; Erismann et al., 1977; Masch et al., 1985), radiometric age dating (Ivy-Ochs et al., 1998), seismic subsurface investigations (Brückl et al., 2001), geochemical characteristics of the slide deposits (Purtscheller et al., 1995, 1997) and kinematical considerations (Erismann and Abele, 2001; Sørensen and Bauer, 2003).

In contrast to the well-investigated accumulation area, the source area of the Köfels rockslide has not been a primary focus of geological surveys yet. Although the spatial distribution of lithological units was mapped already formerly (Hammer, 1929; Sieder and Pirchl, 1994), the structural characteristics of the scarp and their relation to slope failure mechanisms still posed fundamental questions about the kinematical processes involved. Therefore, new field surveys of the scarp and deposits were carried out in order to study the influence of lithologies and geological structures on rock mass failure. Presenting the geological constraints of the Köfels rockslide, this study aims to contribute to an improved knowledge about structurally controlled failure and slope deformation behaviour in metamorphic bedrock units.

2. Geological setting

The Köfels rockslide is situated in the central Ötz valley (Eastern Alps, Tyrol), a NNW-trending main tributary of the Inn valley, where at least 400 m thick slide deposits separate the striking Längenfeld basin in the south and the Tumpen basin in the north (Fig. 1). Surrounded by summits exceeding elevations of 3.000 m a.s.l., this area is deeply incised in the polymetamorphic Ötztal complex, a major thrust unit belonging to the Upper Austroalpine basement nappes. Lithologically a variety of metapelites and metapsammites (paragneisses, micaschists) is encountered, with intercalations of acid to intermediary metamagmatites (orthogneisses), amphibolites and eclogites (Hammer, 1929; Purtscheller, 1978). Locally these rocks were discordantly intruded by numerous post-Variscic and pre-Mesozoic diabase dikes of basaltic-andesitic composition (Purtscheller and Rammlmair, 1982).

The complex structural setting of the polyphase and heteroaxially deformed Ötztal basement may be attributed to at least three distinct orogeneses and their corresponding regional metamorphisms, belonging to the Caledonian, Variscian and Eoalpine phases (Sassi et al., 2004). Firstly, radiometric dating data derived from several orthogneisses range between 470-420 Ma, indicating according ages of the protoliths and metamorphosis respectively (Hoinkes and Thöni, 1993; and references therein). Secondly, the main mineral paragenesis, fabrics and geological structures are to be attributed to a dominant Variscian metamorphisms, dated at about 270-300 Ma and featuring an amphibolite-facial maximum in the northern Ötztal Alps. Thirdly, in the Cretaceous, a somewhat lower graded Alpine regional metamorphism in amphibolite facies affected the Ötztal basement, increasingly from NW to SE with a maximum in the southern Ötztal Alps (Hoinkes et al., 1982). The main geological structures, comprising E-W trending and other large-scale fold systems, were formed by at least four ductile deformation phases of Paleozoic to Alpidic (Cretace-ous) ages (Van Gool et al., 1987).

In contrast to numerous petrological and geochronological studies, the brittle deformation of the Ötztal basement was not investigated systematically so far. However, some information may be provided indirectly by structural analyses of adjacent geological units. As a result of W- to NW-directed Alpine thrust and strike-slip tectonics, the northern Ötztal unit borders along the generally steeply inclined Inntal fault zone to the Alpidic nappe stack of the Northern Calcareous Alps. There, a complex pattern of deformation zones, including E-W trending fold- and fracture systems as well as NE-SW trending sinistral and NW-SE trending dextral major faults, is to be differentiated (Eisbacher and Brandner, 1995; Ortner, 2003). Along the NE-SW-trending Engadin line in the west and the N-S-trending Brenner fault in the east, the Ötztal basement borders on Penninic and Lower Austroalpine basement nappes (Brandner, 1980; Schmid et al., 2004).

In the Quaternary, the Ötz valley region was affected by repeated glacier fluctuations, causing distinct sediment accumulation as well as glacial and fluvial erosion (Senarclens-Grancy, 1958). Fault-related valley deepening, which uncovered preferentially orientated sliding planes and caused substantial stress redistributions in the steepened slopes, is certainly a main factor contributing to rock slope failures in Alpine environments. Compiled radiometric dating data show that the majority of dated landslides in the Eastern Alps, including some major events in the Ötz valley, did not fail immediately after late-Pleistocene glacier retreat and unloading but occurred some 1.000 years later in the Holocene after complex and time-dependent processes of crack growth and fracture propagation (Prager et al., 2008). However, the Quaternary filling of the Ötz valley is characterised by significant valley steps and flat upstream valley floors that may be attributed to multiphase landslide events and therewith associated backwater sediments. Here the major morphological features are the genetically complex landslide-dammed Tumpen basin (Poscher and Patzelt, 2000), and the fluvio-lacustrine backwater-deposits in the Längenfeld and Niederthai areas (Fig. 1), where both the main Ötz and the tributary Horlach valley were blocked by the Köfels rockslide (Ampferer, 1939; Klebelsberg, 1951; Heuberger, 1975).

3. Methods

Detail lithological and structural mapping led to a comprehensive geological map of the scarp and accumulation area of the Köfels rockslide (Fig. 2). The geological field data were completed by analysing optical colour images and airborne laserscanner images (Tiris, 2009) as well as geological cross-sections. The topographic surface lines of latter were derived from high-resolution airborne laserscanner data (LiDAR), with an accuracy of at least 1 measurement point per 1 m² at elevations up to 2000 m asl and 1 point / 4m² above 2000 m asl respectively (Amt der Tiroler Landesregierung, flyover 2006), and yielded valuable information concerning the morphological and structural characteristics of the rock slopes deformation behaviour.

In order to investigate potentially active slope deformations in the Köfels area, satellite based repeat pass radar images (Interferometric Synthetic Aperture Radar, InSAR) acquired within a time period of 1 year were analysed. Across-track SAR interferometry is based on the detection of phase differences between two SAR images acquired at different dates from similar orbital positions (repeat pass observation) and enables the measurement of the displacement component in the line of sight direction of the radar system with very high accuracy (e.g. Rott et al., 1999). The method, capabilities and limitations of SAR interferometry for detecting and monitoring slope motion in mountainous areas are described in Rott and Nagler (2006). For studying the Köfels rockslide, a SAR image pair acquired by the Advanced Synthetic Aperture Radar (ASAR) onboard of ENVISAT during ascending passes

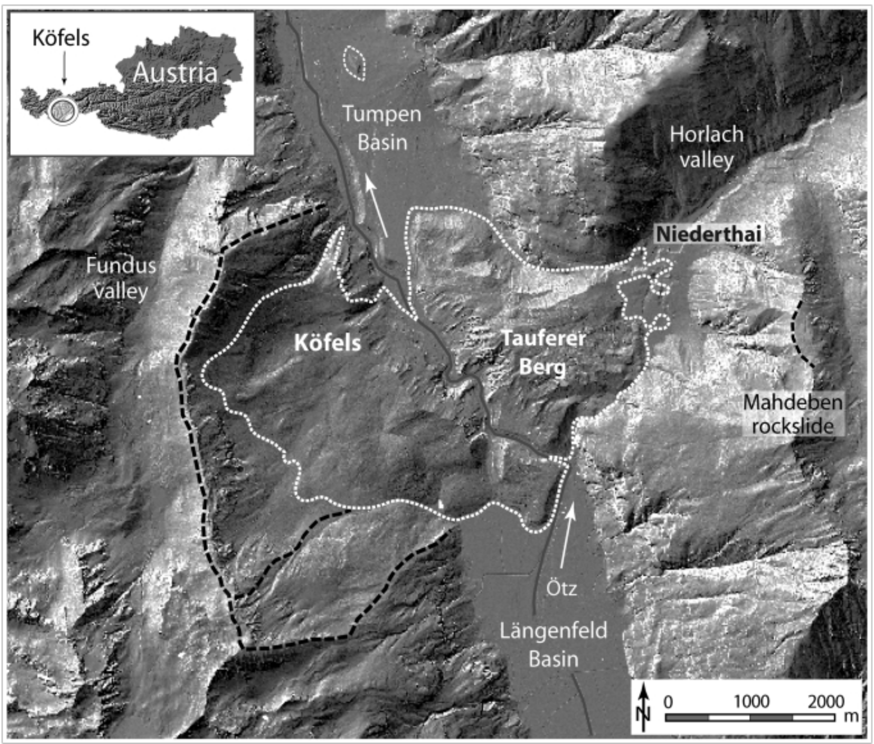

Figure 1: Shaded relief map (airborne laserscanner image; Tiris, 2009) showing the scarp (black stippled) and accumulation area (white stippled) of the Köfels rockslide, as well as the backwater areas in the Ötz valley (Tumpen and Längenfeld basin) and Horlach valley (Niederthai area). Note isolated Toma hill in the Tumpen basin (location Lärchenbühel).

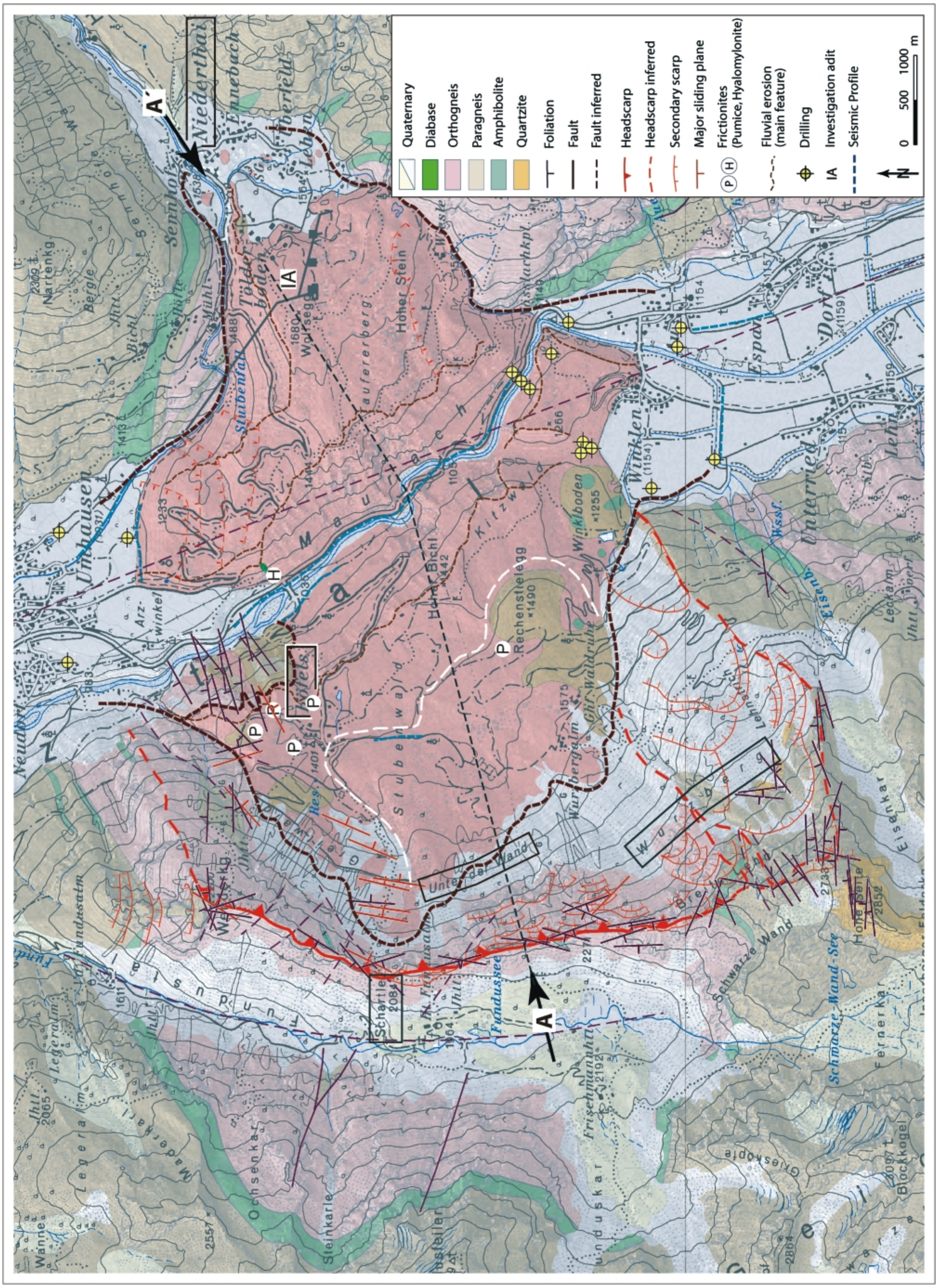

Figure 2: Geological map of the Köfels rockslide area. Frictionites (Preuss, 1974; Masch et al., 1985), drillings (Klebelsberg, 1951; Brückl et al., 2001), investigation adit (IA) Horlach valley, i.e. place of finding of C-14 dated wood fragments buried by the rockslide deposits (Heuberger, 1966) and seismic lines (Brückl et al., 2001) depicted according to the references cited. Displaced rockslide masses enveloped stippled dark red; inferred secondary failure event southwest of Köfels (white stippled) according to Hermanns et al. (2006) and own observations. A-A´: cross-section (Figs. 9, 10).

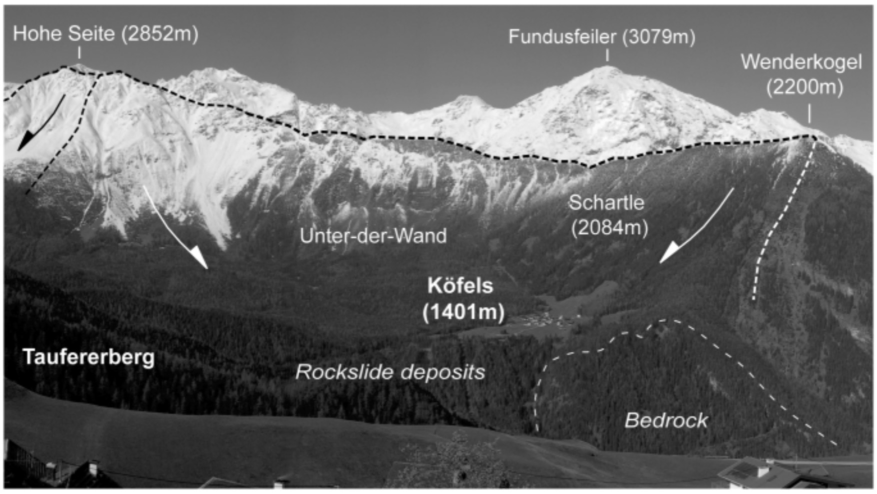

Figure 3: Scarp and proximal accumulation area of the rapid Köfels rockslide (view from Höfle/Niederthai towards SW). Indicated are in-situ bedrocks below Köfels, slope collapse between Wenderkogel and Hohe Seite (white arrows) and slowly creeping paragneiss rockslide (black arrow) at the Hohe Seite area (cf. Fig. 2).

(flight direction from N to S) on 7 August 2004 and 27 August 2005 was available. ASAR operates at a frequency of 5.6 GHz (C-Band), and has a repeat pass observation interval of 35 days. The steep topography in the investigation area causes strong distortions in the radar image. As the Köfels rockslide is oriented towards east, it is observed with higher resolution in the ascending radar images.

4. Accumulation area

The Köfels rockslide detached along an east-facing slope from the mountain ridge between Wenderkogel (2200 m) and Pt. 2733 m (near Hohe Seite, 2852 m), leaving a giant amphitheatre-shaped niche (Figs. 1-3). The accumulated debris, covering at least 11.5 km², blocked both the main Ötz valley and the opposing mouth of the W-trending tributary Horlach valley, making up the huge debris ridge of the Taufererberg (Wolfsegg 1680 m; Figs. 1, 4).

The present surface of the rockslide deposits is widely covered by forest (mainly *Picea*) and scrub, but characterised by several ridges and dome-like elevations with associated depressions in between. Toma, i.e. cone-shaped hills made up by rockslide debris and being typical for the nearby carbonate rock avalanche deposits of Tschirgant and Fernpass (Abele, 1974; Prager et al., 2008, 2009), are encountered decametre-sized in the distal accumulation areas only. For example in the terrace scenery around Niederthai (1538 m) and presumably also in the Tumpen basin at Lärchenbühel (approx. 1030 m), where an isolated hill of coarse (ortho-)gneiss debris occurs in between on-lapping debris flow deposits (Fig. 1; Hammer, 1929; Senarclens-Grancy, 1958). However, neither the unsmoothed morphology of the hilly accumulation area nor the rough scarp show any signs of glacial overprints and indicate a post-glacial genesis of the hummocky rockslide scenery. These field evidences are consistent with radiocarbon dating of buried wood and surface exposure dating of rockslide boulders, according to which a major slide event occurred in the early Holocene at about 9800 yrs (Ivy-Ochs et al., 1998). Geological field surveys show that the majority of the naturally exposed rockslide deposits are made up by fragmented orthogneisses (granitic augen- and flasergneisses), i.e. cubic to cuboid blocks of some decimetres to several metres in diameter. Forming clast-supported fabrics, the pore-spaces are filled with stony to sandy rock fragments but are often devoid of finer grained matrix in near-surface outcrops. Indicated by the absence of springs and significant surface discharge, as well as by the occurrence of noticeable airflows within the debris, these successions are characterised by a high porosity and permeability respectively. Some marginal sections of the proximal deposits, e.g. around the hamlet Köfels in the north and around the morphologically characteristic Rechenstielegg (1490 m) ridge in the south, are composed of paragneisses with some amphibolites and quartzites involved (Fig. 2). Due to their generally lower rock strength and the occurrence of densely spaced discontinuities, i.e. layering, foliation and fracture planes, the incompetent paragneiss series typically disintegrated into cuboid to platy rockslide boulders, generally smaller in diameter and more densely vegetated than the competent orthogneisses. At several locations, especially at Taufererberg, the angular rockslide debris is associated with glacial, fluvio-glacial and fluviatile deposits featuring well-rounded clasts. Presumably these gravels were scraped from the paleo-valley floor and transported by the rockslide masses up to their present position (Ivy-Ochs et al., 1998, and references therein; Hermanns et al., 2006). In contrast, some pebbles of fluvio-glacial origin around Köfels, i.e. situated on top of the proximal to medial rockslide deposits, can genetically not be derived from the valley floor; rather these relict Pleistocene sediments must have been transported from the scarp area, piggy-back on top of the failing rock slope, similar to the situation at the nearby rock avalanches at Fernpass and Tschirgant (Prager et al., 2009).

The internal structure of the Köfels deposits is characterised by heterogeneous block-size distributions and by spatially strongly varying degrees of fragmentation, resulting in laterally and vertically differing types of fabrics. Depending on the accumulation position, rock fragments of different sizes, i.e. from gravels to blocks, and different degrees of fragmentation are embedded in a densely packed unsorted matrix, the grain-sizes of which range from silty-sandy to stony-blocky. The unsorted debris masses, especially the exposed topmost successions, show a chaotic clast-supported fabric with a fractal-like block-size distribution and typically contain crushed rockslide boulders with a jig-saw-fit of corresponding grain boundaries. Lower sections of the thick Köfels deposits, but not their basal units or their substrate, are naturally exposed along the approx. 3.5 km long and approx. 400-500 m deep fluvial canyon “Maurach” (Figs. 1, 2, 4). Here penetrative fragmentised and finely ground rocks are encountered, but also some relatively compact rock slabs of several decametres in diameter, which here and even higher up at Taufererberg have formerly been mapped in parts as in-situ bedrocks (e. g. Hammer, 1929; see also Sørensen and Bauer, 2003). In the northern Maurach a several metres large diabase dyke, which embedded in its surrounding granitic-gneiss wall rocks was transported in the course of the rockslide, shows only a crushed and heavily fractured texture but no features of remarkable shear deformation along the intrusion contacts.

One outstanding characteristic of the Köfels site consists of the existence of fused rocks, which, according to their texture and assumed genesis, have been referred to as “pumice“ (Pichler, 1863; Preuss, 1974), “Köfelsit” (Suess, 1937), “impactites” (Kurat and Richter, 1972), “frictionites” (Erismann et al., 1977) and “hyalomylonites” (Masch et al., 1985). So far, landslide-related rock fusion has only been reported from four other landslides worldwide, namely from the base of Tsergo Ri/Nepal, Arequipa/Peru, Dzongri/India (Weidinger and Korup, 2009), all with failure volumes > 1km^3, and from a smaller (3x10^7 m^3) loess-covered mudstone slide at Sale Mt./China (Zhang et al., 2002). According to the studies cited, the frictional melting of the Köfels (ortho-)gneisses along distinct sliding planes resulted from very high strain rates due to the volume and velocity of the rockslide mass. Field outcrops exhibit two macroscopic main types of dynamically deformed and fused rocks, depending on their degree of degasification, i.e. porous frictionites (“pumice”, thickness ranging from cm to several dm) in the Köfels area (at ca. 1400 m asl) and dense glasses (“hyalomylonite”, thickness mm to some cm) along the Maurach gorge (at ca.1050 m asl) (Figs. 2, 4). Resulting from the dynamic disintegration and fragmentation, locally radioactive springs discharge and very high radon gas concentrations are emitted from the highly fractured and crushed rockslide deposits (Sieder and Pirchl, 1994; Purtscheller et al., 1995, 1997).

Another striking feature of the Köfels rockslide mass is the occurrence of widely spread rockslide-dammed backwater-deposits (Fig. 1). In the southerly adjacent Längenfeld basin, the extensive geological subsurface investigations for an abandoned hydropower project showed that the fluvio-lacustrine deposition sequences accumulated here up to at least 135 m thick, comprising fine-grained backwater sediments of up to 92 m in thickness. Reflection seismic measurements yielded evidence that the top of the compact bedrock units plunges here from approx. 50-135 m below ground in the Längenfeld basin steeply northwards to approx. 400 m below ground at the toe of the Köfels slope and thus make up a significant paleo-valley step of about 300 m (Klebelsberg, 1951; Brückl et al., 2001). Furthermore, hydrogeological field tests indicate that the thick Köfels rockslide deposits are characterised by relatively high hydraulic permeabilities; therefore the river Ötz formerly discharged most likely subterraneously through the damming rockslide barrier and the level of the accumulated backwater-deposits did not exceed the present-day surface of the Längenfeld basin (Klebelsberg, 1951; Heuberger, 1994). In the easterly adjacent Niederthai area, the terrace sediments are presumably a few tens of metres thick, but to which amount these are to be attributed to a periglacial genesis (Senarclens-Grancy, 1958; Heuberger, 1966) and/or to the Köfels event (i.e. rockslide dammed or mobilised valley floor deposits; see Geitner, 1999, p.17), has not been solved clearly yet. In the north, the Köfels deposits were transgressed by fluvio-lacustrine backwater deposits, with thicknesses of > 150 m due to blockages by rockslide and rockfall events in the Tumpen area (Poscher and Patzelt, 2000).

5. Scarp area

The well-exposed head scarp of the Köfels rockslide is characterised by an east-facing amphitheatre-shaped niche (Figs. 1-3). Its western boundary is clearly traceable along the sharp mountain ridge, extending from Wenderkogel along the incision Schartle towards Hohe Seite. The northern scarp flank is morphologically less evident, but situated somewhere in the steep and rugged rock slope northeast of Wenderkogel. The southern boundary of the Köfels niche was previously not clearly defined, but was depicted in a few maps to extend from Pt. 2733 (northeast of Hohe Seite) down slope along a broad east- to northeast-trending ridge (Brückl et al., 2001; Hermanns et al., 2006). This lateral boundary is of major importance for the geological slope model, failure mechanisms

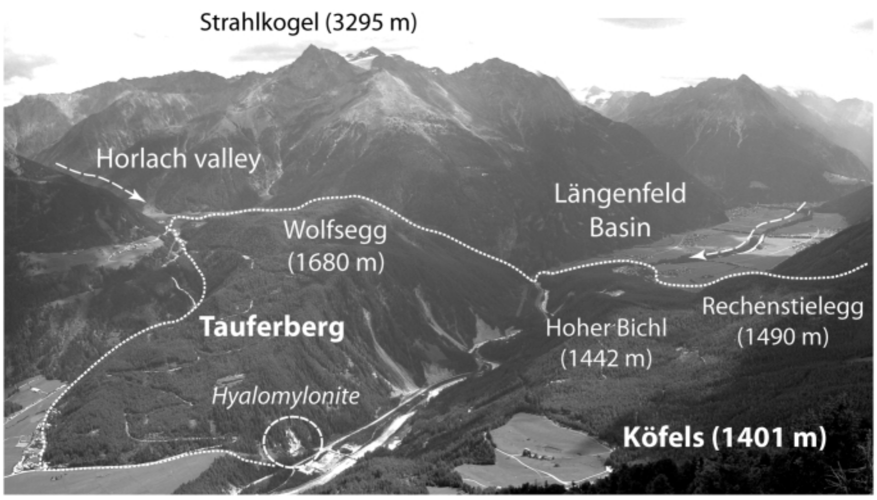

Figure 4: Main accumulation area of the Köfels rockslide showing several hundred metres thick deposits and therein incised river Ötz with rockslide-dammed Längenfeld basin in the background.

and accumulation scenarios of the Köfels rockslide and thus is described in the context of this study.

5.1 Lithological predisposition

In the Köfels area the slope bedrock is comprised of three major lithological and structural bedrock units, all several hundred metres thick, which are i) a paragneiss series including amphibolites in the Wenderkogel area and further north, ii) a granitic orthogneiss complex in the northern to central scarp area, extending from Wenderkogel in the north to Wurzberg in the south and iii) another paragneiss series including amphibolites and quartzites, extending from Wurzberg to Hohe Seite in the south (Fig. 2).

According to the lithologies encountered in the scarp and accumulation area, the orthogneiss complex surrounding the hamlet of Köfels clearly represents the main source area of the Köfels rockslide. The massive to weakly foliated granitic augen- and flasergneisses are characterised by cm-large feldspar-eyes and -flasers that are embedded in a matrix of quartz, feldspar, white-mica and biotite. These and the mineralogical equivalent granite-gneisses, i.e. orthogneisses without striking feldspar-eyes, do not show geochemical peculiarities except some slightly elevated uranium contents (Purtscheller et al., 1995, 1997). In both the north and south the firm orthogneisses border to thick successions of closely foliated paragneiss series that make up the southern scarp niche, i.e. the area Wurzberg to Hohe Seite. Petrographically these para-rocks comprise biotite-plagioclas-gneisses, macroscopically similar and thereof hardly differentiable mineral-rich micaschists (Sieder and Pirchl, 1994) and a few tens of metres thick layers of quartzites and amphibolites, with the latter mainly occurring near the contact to the orthogneisses. In the northern scarp area, some paragneisses are intercalated in the orthogneisses, e.g. along the path from Köfels up to Wenderkogel, and alternating with each other respectively, such as at the steep slope declining from northeast of Köfels down to the river Ötz. The latter units either represent extraordinary large but only minor displaced (“parautochthon”) sliding slabs or even in-situ bedrocks (Figs. 2, 3); anyway these are not part of the “typical” rockslide deposits, but rather make up a zone of low rock mass strength at the toe of the slope and thus might have favoured the slope failure. Potential “markers” for rockslide displacements are given by fine crystalline and only a few metres thick diabase dykes, which locally intruded both the ortho- and para-rocks along the Wenderkogel - Hohe Seite ridge.

5.2 Tectonic structures

The structural setting of the Köfels area is characterised by a striking orthogneiss complex embedded in surrounding paragneiss series. The spatial orientation of the layering and main foliation of the orthogneisses (Fig. 5c, Tab. 1) clearly indicates an open first-order antiform (structural dome) with a moderately W-plunging b-axis (mean orientation: Eigenvector 247/18). Therefore the orthogneisses are i) overlain and enveloped by paragneiss series in the scarp area and west of

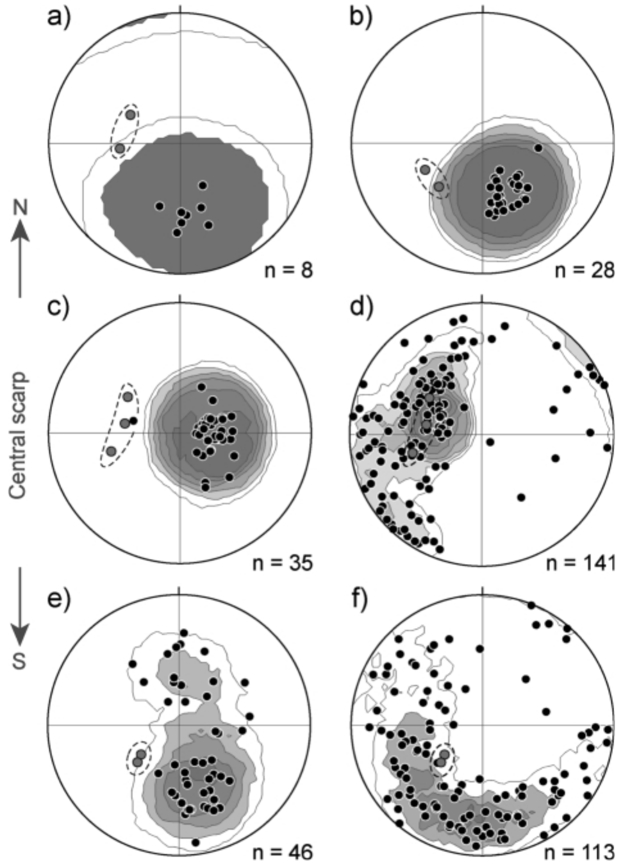

Figure 5: Measured discontinuities (black, poles to planes, lower-hemisphere) and mean slope orientation (grey, poles to planes, stippled enveloped) of the Köfels rockslide scarp area: foliation of a) paragneisses in the northern scarp area (Wenderkogel), b) para- and orthogneisses at the slope northeast of Köfels, c) orthogneisses in the central scarp area and e) paragneisses in the southern scarp area (Hohe Seite), as well as fractures of d) orthogneisses in the central scarp area and f) paragneisses in the southern scarp area.

the Fundus valley, ii) obviously tightly folded and/or faulted east of the Ötz valley and iii) terminating around the meridian of Niederthai (Hammer, 1929; Sieder and Pirchl, 1994). According to the geological map, the northern contact to the overlying N-dipping paragneisses (Figs. 2, 5a, b) is rather conform in the Wenderkogel area, whereas the southern contact in the area of Wurzberg - Hohe Seite is more complex: here the W- to SW-dipping orthogneisses (and amphibolites) border, along an approximately N-dipping contact zone, to a generally N-dipping paragneiss series; the geometry of this unconformity may be attributed to a primary intrusion contact and/or to post-intrusive tectonic deformations. How far the othogneisses of the prefailure topography originally extended further to the south/ southeast, i.e. to which extent they overlaid the adjacent paragneisses of the Wurzberg area, can hardly be estimated. These paragneisses are generally E-W-trending (Fig. 5e, Tab. 1) and, at several locations, internally folded in the metre to 100-m scale. Especially in the Hohe Seite area, the tight asymmetric subsidiary folds feature N-dipping long fold limbs and S-dipping short limbs, with moderately W-plunging b-axes (mean orientation: Eigenvector 263/11).

Concerning the brittle tectonic deformations, different sets of meso-scale tensile joints and shear fractures are encountered in both the ortho- and paragneissic bedrocks. Several discontinuities are coated with chlorite, a few only with quartz. Fault gouges, kakirites or noticeable fault breccias have not been observed so far. The collected field data from the orthogneisses (og) at the SE- to E-dipping slope of the northern and central scarp areas show a scattered spatial distribution (girdle distribution). However, based on overall geological field observations and statistical cluster analyses, four main fracture sets were differentiated (Fig. 5, Tab. 1).

Fracture set og#1 comprises the most striking structural features of the Köfels site, the well-known major sliding planes that are exposed along the path to Schartle (Figs. 2, 6) and forest roads nearby. Here these rather gently inclined key-fractures, featuring a mean orientation (azimuth/dip) of approx. 100/30 and thus day-lighting sub-parallel to the east-facing slope, show remarkable planar surfaces (Preuss, 1986) and trace lengths of up to several tens of metres. Some of these weakness planes are slickenside coated with quartz, a few exhibiting thin cataclasites (attrition breccias) and others featuring unpolished i.e. rather rough surfaces. Commonly they are devoid of significant striations, but polished and thus showing evidence for shear deformation, most likely of tectonic origin (see Sect. 7) and/or due to the rockslide event. South of the Schartle - Köfels area, the moderately dipping fractures og#1 obviously decrease in size, density and frequency of occurrence. Other brittle major fault zones of comparable spatial extent, and causing such distinct rock strength anisotropy, are not encountered in the scarp area. Fracture set og#2 comprises steep N-S-striking planes and may be associated with the deeply incised N-trending Fundus valley west of the Köfels scarp (Figs. 1, 2). Central sections of the Köfels head scarp, especially the subvertical rock walls at the location Unter-der-Wand (Fig. 7), exhibit relatively intact orthogneisses that are cut by approx. 60-80° E-dipping main fracture zones. Posing preferred zones of rock weakness, these discontinuities favoured the observed gravitational slope deformations (see Sects. 6, 7). Fracture set og#3, i.e. steeply NW-trending discontinuities, is obviously orientated sub-parallel to the central Ötz valley and corresponds to the present-day stress field, with amaximum principal horizontal stress orientated NNW-SSE (Drimmel, 1980; Heidbach et al., 2008). This and the tempered spring "Bad Längenfeld" (southern Längenfeld basin, cf. Fig. 1), which emanates from a NW-trending joint (Zötl and Goldbrunner, 1993), indicate deep-seated flow systems and according permeable main fracture zones. Fracture set og#4 comprises steeply NE-trending discontinuities (plus random distribution), which genetically may be attributed to some minor fault zones occurring e. g. in the south-eastern scarp region (see below).

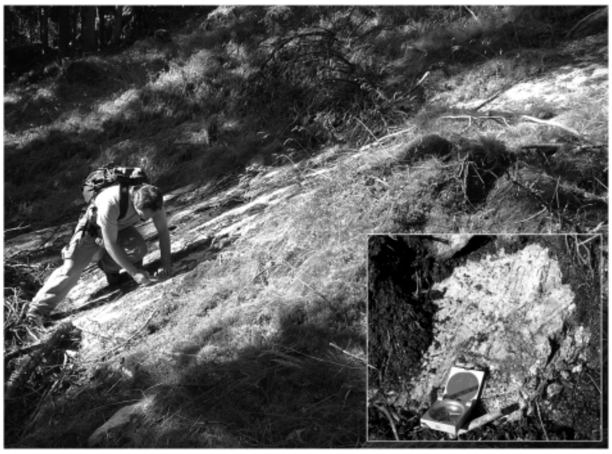

Figure 6: In-situ orthogneiss with gently E-dipping (mean azimuth/dip: 100/30), polished major sliding plane of the Köfels rockslide (between Köfels and Schartle, at approx. 1800 m a.s.l.).

Similar to these findings, also the fractures measured in paragneisses (pg) at the southern scarp region yielded a substantial spatial scatter. The pi-plots show a rather broad girdle distribution, i.e. these data are not clearly attributable to individual clusters with representative mean orientations, but based on the field observations four main fracture sets were differentiated (Fig. 5f, Tab. 1). Set pg#1 is characterised by a cluster of moderately E-dipping meso-scale fractures. Fracture set pg#2 comprises some main brittle discontinuities orientated subparallel to the predominantly N-dipping layering and main foliation. Set pg#3, i.e. steeply NW-trending fractures, may also form a girdle distribution with set pg#1. Set pg#4 is mainly represented by a NE-SW-trending fault zone southeast of the scarp and thus may be somewhat overem-

Discontinuity set	Orientation (trend)	Statistical mean	n	Remarks
foliation og (Fig. 5c)	NE-SW to N-S to NW-SE	272/20	35	W-plunging antiform
foliation og/pg (Fig. 5b)	E-W to NE-SW	334/35	28	
og#1 (Fig. 5d)	N-S	97/32	57	major sliding planes
og#2 (Fig. 5d)	N-S	90/76	23	head-scarp deformation
og#3 (Fig. 5d)	NW-SE	47/83	32	orientation Ötz valley
og#4 (Fig. 5d)	NE-SW	152/52	29	
foliation pg (Fig. 5a)	E-W	354/46	8	
foliation pg (Fig. 5e)	E-W	341/42 (and 187/34)	46	W-plunging fold axes
pg#1 (Fig. 5f)	N-S	104/41	31	
pg#2 (Fig. 5f)	E-W	356/61	30	
pg#3 (Fig. 5f)	NW-SE	40/70	27	orientation Ötz valley
pg#4 (Fig. 5f)	NE-SW	300/77	25	

Table 1: Compiled mean orientations of discontinuities measured in ortho- (og) and paragneisses (pg) at the Köfels slope. Foliation differentiated for areas Schartle - Unter-der-Wand (Fig. 5c), bedrock slope northeast of Köfels (Fig. 5b), Wenderkogel (Fig. 5a), Hohe Seite – Wurzberg (Fig. 5e). Statistical mean values (azimuth/dip) of foliation and fracture sets based on cluster analyses (Giné statistic; see also Fig. 5).

phasised in the depicted data. However, in the paragneisses of the southern scarp area several out-of slope discontinuities are encountered, but no evidence of major sliding planes dipping down slope as observed in the northerly adjacent orthogneisses (i.e. set og#1).

6. Slope deformations

The geological setting and morphological features of the scarp area point to different types of failure mechanisms. In general, to be differentiated are i) the prominent Köfels rockslide event and ii) ongoing instabilities affecting the steep post-failure orthogneiss slope (Unter-der-Wand area) as well as iii) a slowly creeping rockslide in the adjacent paragneisses (Wurzberg area).

6.1 Pre-historic rockslides

Radiocarbon dating of buried wood, found in the investigation adit Horlach valley (Fig. 2), and surface exposure dating of rockslide boulders indicate that the Köfels rockslide event occurred in the early Holocene at about 9800 yrs and was succeeded by at least one minor event (Ivy-Ochs et al., 1998; Kubik et al., 1998, 2007). In view of these data and morphological features, Hermanns et al. (2006) roughly differentiated between two slide masses: firstly a major event accumulating in the Taufererberg area, and secondly a clearly smaller event in the hummocky scenery south of Köfels. In how far multiple slope failures can here be backed up by mapping the spatial distribution of ortho- and paragneissic rockslide debris (Fig. 2) and/or different fragmentation patterns, has not yet been solved satisfyingly, but poses a challenging topic for future investigations. However, the radiometric dating data indicate that the failure of the fluvio-glacially steepened Köfels slope was evidently not directly triggered by deglaciation processes, but took a preparation time of some 1.000 yrs after retreat of the late-Pleistocene glaciers. Basically this slope collapse can be attributed to its lithological, structural and morphological predisposition as well as to time-dependent rock strength degrading processes of crack growth, fracture propagation and coalescence of brittle discontinuities.

The geological field surveys performed at Köfels show that the NW-, W- to SW-dipping foliation planes of the orthogneisses, i.e. the main source of the Köfels rockslide, are orientated against the E-facing slope and thus unfavourable to promote deep-seated slope instabilities (Figs. 2, 9). In contrast, the paragneisses show mainly N-dipping foliation planes, which provided - in the southern scarp area - potentially weak detachment zones for the adjacent orthogneisses. These layering conditions and the morphological-structural characteristics of the well-exposed scarp indicate that the progressive slope deformation and final failure were clearly structurally predisposed and controlled by brittle fracture systems. Here the key-discontinuities that reduced the strength of the orthogneisses and highly influenced rock mass anisotropy are two main sets of differentially inclined E-dipping large- and mesoscale fractures (joints, faults). Staircase-like slope profiles (Fig. 10) and several outcrops exhibiting stepped rock surfaces show that ca. 25-35° E-dipping major fractures, which daylight on the E-facing slope and form preferred sliding planes (set og#1; Fig. 6), are cross-linked by ca. 60-80° E-dipping (set og#2) and other discontinuities. The coalescence of this fracture network caused bedrock fragmentation into major slabs and slide blocks, generated persistent failure planes and finally determined the block size distribution of the slide debris. The lateral boundaries of the Köfels rockslide scarp were predisposed primary by the spatial extent of the orthogneisses and its surrounding E-W-trending paragneisses as well as secondary by NE- and NW-trending fracture sets (og#3-4, pg+#3-4), plus random.

Besides the structural setting, this slope collapse was also lithologically favoured by weak bedrock units exposed at the toe of the slope. According to the geological field surveys, series of para- and orthogneisses are also encountered along the slope northeast of Köfels and covered by rockslide debris further south (see Fig. 2). Azimuth and dip of these jointed but relatively constant 30-45° N- to NW-dipping bedrocks are in accord with those of adjacent in-situ rocks at the scarp (Fig. 5b; Tab. 1). Thus these paragneisses are not part of the rockslide deposits (as depicted e.g. by Brückl et al., 2001; Hermanns et al., 2006), but make up a zone of low rock mass strength at the toe of the slope and might have substantially favoured the failure.

However, the failing rock masses crossed the Ötz valley, hit their opposite slope and were piled up as a remarkably thick debris ridge, which makes up the Taufererberg (Figs. 1, 2, 4, 9). According to the mechanical model by Erismann et al. (1977) and Erismann and Abele (2001), due to this impact the rockslide mass was separated in to two main units: a lower one, which was blocked by the opposite slope, and an upper one that continued moving further east till Niederthai; in between, a secondary shear zone featuring the well-known frictionites was postulated. So far, the existence of this secondary sliding plane has been subject to dispute but neither been coherently verified nor falsified (Brückl et al., 2001; Hermanns et al., 2006). Rather the rockslide deposits exposed along the deeply incised fluvial canyon Maurach (see Sect. 4) feature relative homogeneous slopes with constant inclinations and do not yield morphological evidence (e.g. slope breaklines, terrain flattening, vegetation changes, spring lines etc.) for the presence of major subhorizontal internal sliding planes, along which intensely disintegrated and/or fused rocks are expected to occur.

Concerning the slope collapse, the newly compiled geological map shows that the size of the southern scarp area, which is entirely made up by paragneiss series, evidently differs from the volume of according paragneisses in the accumulation area (Fig. 2). Furthermore, the northeast-facing slope of the southern scarp niche, i.e. the Hohe Seite - Wurzberg area, is characterised by i) a stepped slope profile featuring a concave niche at the upper slope sections and a convex slope toe, ii) secondary scarps and counterscarps (best recognisable in the laserscanner images; Tiris, 2009) and iii) intensively fractured and disintegrated bedrocks (i.e. in contrast to intact in-situ bedrocks in the adjacent scarp areas), so that the middle to lower parts of the slope are widely covered with debris and vegetation. Based on these geological field observations and the slope morphology, considerable parts of the scarp niche, i.e. areas made up by incompetent paragneisses, were evidently not involved in the "rapid" Köfels rockslide event, i.e. the orthogneissic deposits encountered in the Köfels and Taufererberg area. Rather these paragneissic slopes have been affected by slowly creeping deformations termed herein as "Wurzberg rockslide" (Figs. 2, 3, 8, 11; see also Sect. 6.2, 6.3). Ist south-eastern boundary is marked by E- to NE-orientated en-echelon arranged double-crested ridges and extensional depressions, several metres deep and up to some 100-m long. The northern boundary of the paragneissic Wurzberg rockslide is not clearly definable yet, but assumed to be lithologically predisposed by those N-dipping contact zone, along which the intensively foliated paragneisses border on the massive orthogneisses further north.

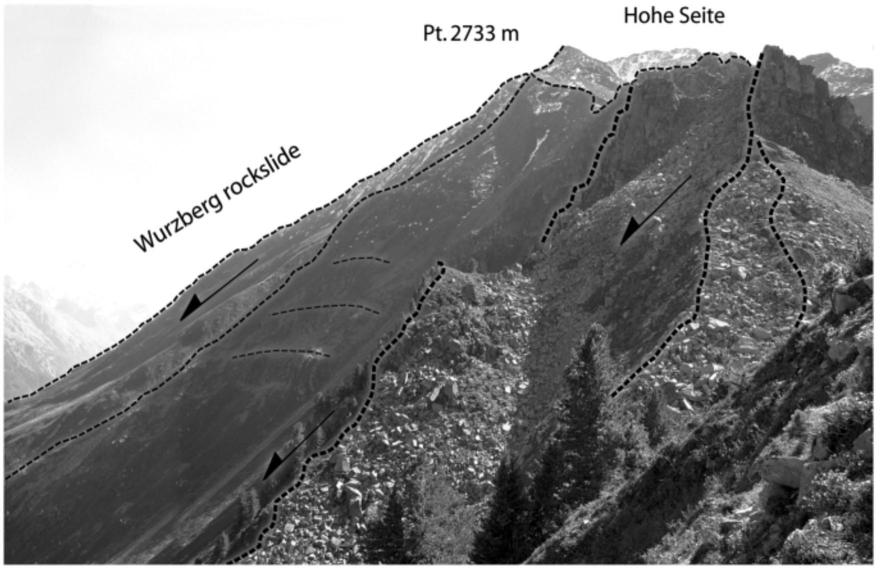

Figure 7: Head scarp of the Köfels rockslide (location: top of Unter-der-Wand, approx. 2200 m a.s.l.) showing post-failure deformations of jointed orthogneisses along 60-80° E-dipping main fracture zones. Bedrock disintegration produces coarse blocky talus deposits and causes some minor rock falls. In the background (left skyline) incompetent paragneiss series affected by the slowly creeping Wurzberg rockslide (cf. Figs. 2, 3, 8, 11).

6.2 Post-failure deformations

The geological field surveys and the optical and laserscanner images (Tiris, 2009) show that the present-day head scarp is widely affected by gravitational post-failure deformations. In northern sections, i.e. at Schartle, N-S-trending double-crested ridges are encountered, with two trench-like depressions in between, which listrically extend >200 m westwards into the fractured in-situ bedrocks. These depressions, >10 m deep and >500 m long, were already described by Hermanns et al. (2006) and interpreted as evidence for ongoing scarp instabilities.

Southerly adjacent, a series of double-crested ridges and en-echelon arranged counterscarps ("Nackentäler") has now been mapped in the well-exposed orthogneisses at the location "Unter-der-Wand". Here the 300-400 m high and steeply inclined (>50° to subvertical) rock walls are cut by several 60-80° E-dipping main fracture zones, which cause E-directed normal faulting of distinct bedrock slabs (Figs. 2, 7, 10). The trace lengths of the individual N-S-trending scarps exceed 500 m, the vertical offsets of the normal-faulted blocks range in the decametre-scale. Evidently the competent orthogneisses deform here along widely i.e. a few decametres spaced main

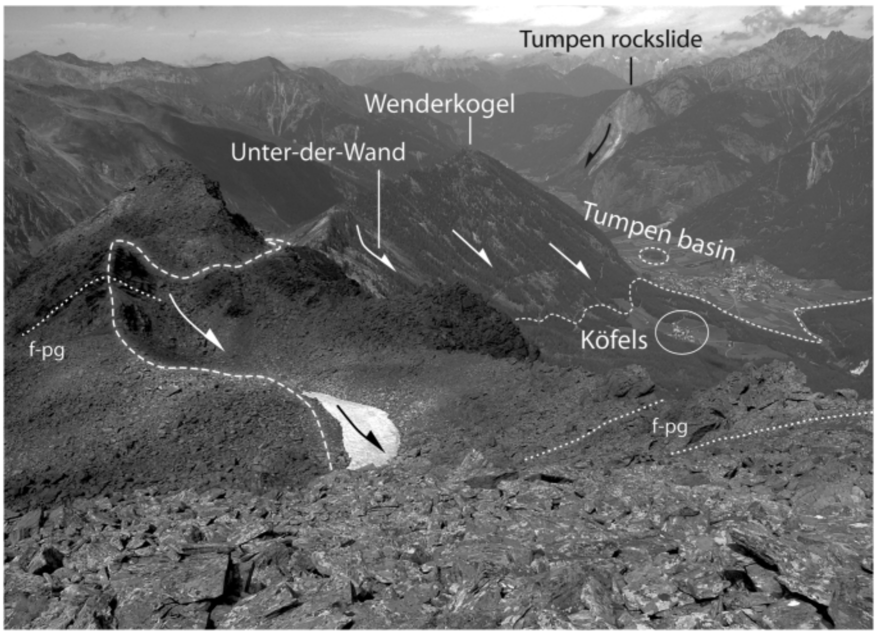

Figure 8: Head scarp of the Wurzberg rockslide (near Hohe Seite, view towards North; cf. Figs. 7, 11): fractured paragneiss series (f-pg = trace of foliation; cf. Figs. 5e, f) featuring deep-seated extension zones and NE-facing secondary scarps, with vertical offsets of around a few decametres. Bedrock widely covered by platy talus deposits, with average block sizes being clearly smaller than those of disintegrated orthogneisses (cf. Fig. 7). In the background, orthogneissic scarp and proximal accumulation area of the rapid Köfels rockslide event, as well as isolated rockslide deposits (Toma hill) in the Tumpen basin (Fig. 1).

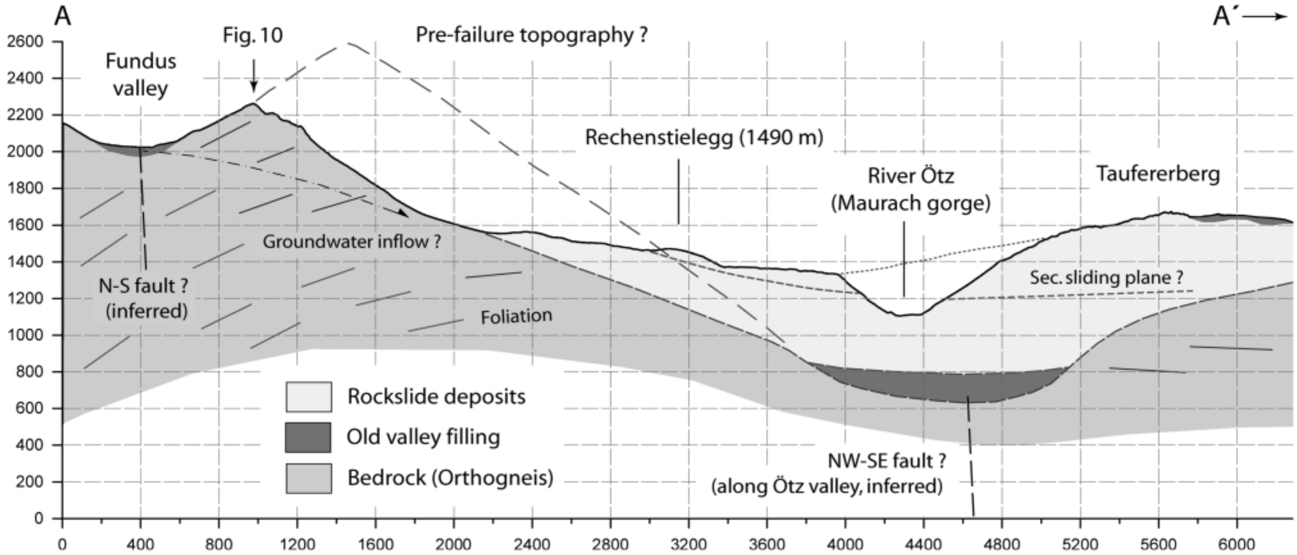

Figure 9: Geological cross section of the Köfels rockslide (cf. Fig. 2), with topographic surface line derived from high-resolution airborne laserscanner data (see Sect. 3). Thickness of slide debris according to Brückl et. al. (2001). Potential groundwater inflows (inferred) from the Fundus valley into the fractured bedrock units schematically indicated.

fracture zones, with relatively intact rock masses in between. In these morphologically striking zones, covered with debris and vegetation, intensely fractured and fragmented fault rocks including finely ground kakirites are expected to occur. However, these findings and the surface topography derived from laserscanning data show that the observed slope breaklines coincide with the orientation of the steeply E-dipping main fracture set og#2 (Figs. 5, 10, Tab. 1). These structurally predisposed main discontinuities pose preferred zones of rock weakness and increased deformability and thus favour the observed slope deformations. As a result, remarkable stepped slope profiles are encountered, featuring i) a distinct head scarp and mountain-splitting in the crest region, ii) several secondary scarps and morphological flattening along the slope and iii) a slightly convex slope toe, where the jointed bedrocks are masked by on-lapping scree and some minor rockfall deposits. These morphological features are generally typical for deep-seated gravitational slope deformations in low strength rocks, e.g. paragneisses, schists and phyllites, but have not been observed in competent bedrock so far. However, the geometry and structure of these slope deformations may be attributed to sliding mechanisms of fractured orthogneisses, controlled by the coalescence of two major fracture systems (i.e. og#1 and og#2), but also to multiple step-path failures and compound failure modes of "rock slumping". Rock slumps are rotational failures occurring in quasi-isotropic weak rocks (Hungr and Evans, 2004), but typically also in fractured hard rock slopes with regular, parallel fractures dipping down slope but not daylighting (i.e. equal to set og#2) and at least one flat-lying fracture that does daylight into free space (i.e. equal to set og#1); the resulting failures are characterised by backward rotations of single or multiple hard rock blocks, which kick out along a basal crossing joint (Goodman and Kieffer, 2000; Kieffer, 2003).

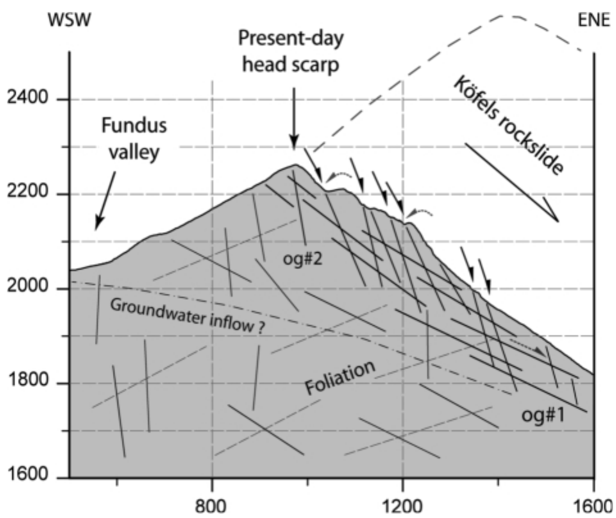

Figure 10: Stepped slope profile encountered in competent orthogneisses (location: Unter-der-Wand), indicating post-failure scarp deformation along en-echelon arranged steep fractures (set og#2; cf. Figs. 5, 7, 11) (i.e. zoom-in of Fig. 9, horizontal scaling according to Fig. 9).

These clearly structurally controlled hard rock slope deformations, observed near Schartle and concentrated in the "Unter-der-Wand" area, continue further south to Pt. 2733m, northeast of Hohe Seite. Here a NNW-SSE-orientated double-crested ridge, featuring a >350 m long and approx. 30-50 m wide distinct major depression ("Nackental", Fig. 8), cuts perpendicular across the foliation of the fractured paragneiss series. This mountain splitting and the geological characteristics of the NE-facing slope, i.e. the occurrence of intensively fractured paragneiss series and the stepped slope profile featuring a scarp niche, a convex slope beneath and secondary scarps, are expression of the deep-seated gravitational slope deformations of the Wurzberg rockslide (Sect. 6.1., 6.3).

6.3 InSAR motion analyses

The geological and morphological field indications of ongoing slope instabilities have been cross-checked with recent analyses of repeat pass satellite-borne radar (InSAR) images. The available InSAR motion field (Fig. 11) is based on the ASAR image pair 7 August 2004 and 27 August 2005. A Digital Elevation Model with 10 m pixel spacing was used for removing the interferometric phase contributions due to topography. Inaccuracies of the DEM can be neglected as the imaging geometry of the image pair has a low sensitivity to elevation, e.g. one interferometric phase cycle of 2π corresponds to an elevation change of 623 m (height of ambiguity). Ano- to an elevation change of 623 m (height of ambiguity). Another error source might be different atmospheric conditions, especially tropospheric water vapour content, at the time of the two image acquisitions. These effects cannot be removed from a single image pair, but are in general at scales larger than several kilometres or tens of kilometres.

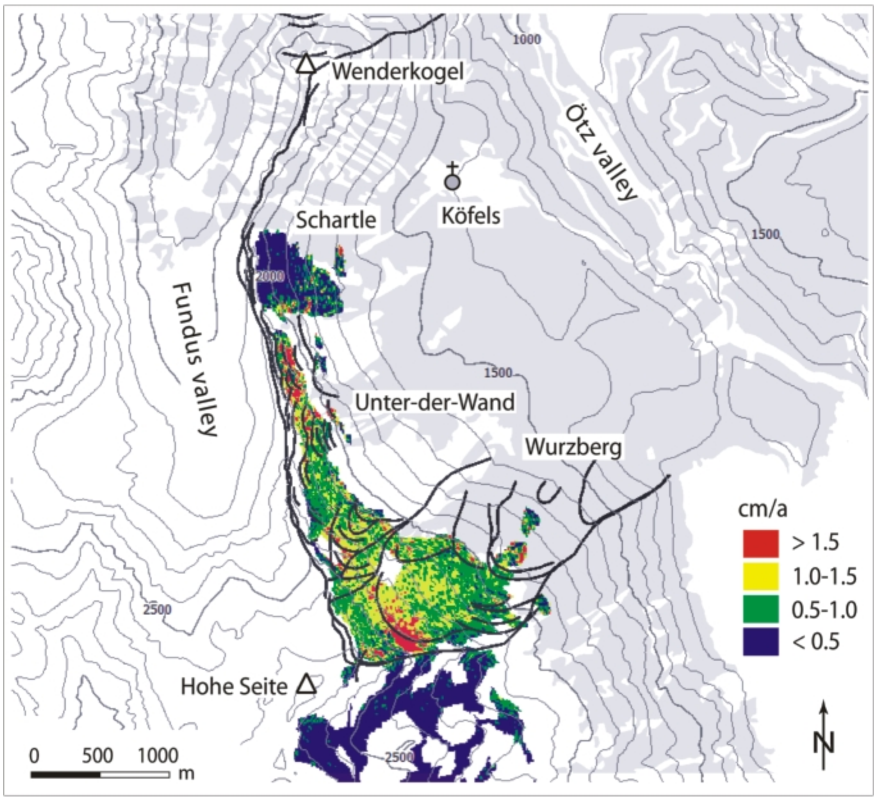

Figure 11: InSAR based motion field (surface parallel assumption) derived from ENVISAT ASAR images (acquired on 7 August 2004 and 27 August 2005) showing active rockslides in the Wurzberg and Unter-der-Wand areas. Scarps (black lines) according to field surveys (Fig. 2). At lower slope sections radar signal decorrelations due to dense vegetation (grey; depicted according to Austrian Map ÖK).

The InSAR based motion analyses show a typical displacement rate of 0.5 to 1.5 cm/a (surface parallel assumption) in the Wurzberg area and in the upper part of Unter-der-Wand. The area below Schartle shows no displacement in the main part, and low displacement rates (< 1 cm/a) in the lower part. At lower sections of the slope (below ca 1800 m) dense vegetation, mainly shrubs and forests, are present, which cause temporal decorrelation of radar signal of the two images permitting the retrieval of motion information (masked in Fig. 11). The size and spatial extent of the motion field indicates, in cross-check with geological field surveys and optical images, that the terrain displacements clearly affect larger bedrock areas. For this study, only one SAR image pair with 1 year time interval was available. The analysis of further SAR image pairs with different time intervals (several months to 2 years) will improve the reliability of the InSAR based motion field, especially when combined with geodetic measurement campaigns, and enables to study the annual velocity variations of the observed rockslides.

7. Discussion and conclusions

The geological and morphological characteristics of the Köfels scarp indicate that the failure of the polyphase and heteroaxially deformed orthogneisses was clearly structurally predisposed by the coalescence of brittle discontinuities. Stability-relevant have been 25-35° E-dipping fault and joint planes, which daylight on the slope and formed preferably oriented sliding planes, and a set of 60-80° E-dipping fractures as well as others. These and the different rock mass strengths of the lithological units encountered controlled the bedrock fragmentation, the block size distribution, the scarp geometry, the sliding planes and finally the slope deformation behaviour.

A comparable structural pattern, i.e. major sets of E-dipping tectonic fractures affecting east-facing bedrock slopes, was also observed at some other instable slopes in the Ötztal basement, e.g. nearby in the Tumpen rockslide area (cf. Poscher and Patzelt, 2000) and at a paragneissic slope southeast of Niederthai (Mahdeben rockslide; see Fig. 1). Also the well-investigated active Hochmais-Atemkopf (Kaunertal) and Steinlehnen (Sellraintal) rockslide systems deform along E-dipping fracture sets (Zangerl et al., 2007). In a regional context, the tectonic origin of these main discontinuities has not been investigated systematically so far.

In the Köfels area, extensive drilling campaigns and reflection seismic measurements yielded evidence that the top of the compact bedrock units plunges from approx. 50-135 m below ground in the Längenfeld basin steeply northwards to approx. 400 m below ground at the paleo-slope toe (Klebelsberg, 1951; Brückl et al., 2001). This significant and presumably fault-related paleo-valley step of about 300 m certainly contributed to the slope collapse by uncovering potential sliding planes and increasing the stresses at the fluvio-glacially steepened slope. These assumptions are in accord with results obtained from numerical modelling of the initial phase of the Köfels rockslide. Isotropic, i.e. disregarding geological structures, 2D models indicate that the maximum shear stress concentrations occurred at the toe of the pre-failure slope (Brückl et al., 2001; Hermanns et al., 2006). Now the geological field mapping suggests that the Köfels rockslide probably was also favoured by the occurrence of incompetent paragneisses at the slope toe. These intensively foliated and jointed bedrocks feature generally low shear strengths and thus might have provided potential detachment zones for the adjacent orthogneisses. However, after initial phases of rock strength degradation and primary and secondary creeping processes and after overcoming the cohesion of the sliding planes, the Köfels slope accelerated until an abrupt final collapse set in (Brückl, 2001). Based on this, the progressive failure was initiated by long-term preparatory stages, in which gravitational creep and tension fracturing led to coalescence of brittle discontinuities ending with shearing along continuous sliding planes.

According to the model of progressive failure, fully persistent discontinuities are rarely encountered and slope failures are much rather induced by the coalescence of discontinuities due to the propagation of pre-existing fractures as well as growth of new ones (Einstein et al., 1983; Eberhardt et al., 2004). Fracture propagation and the stability of rock slopes are generally determined by the size, orientation and density of fractures and strongly influenced by the existing stress field (Einstein, 1993; Einstein and Stephansson, 2000). However, there are complex physical-chemical processes in fractures that enable slow crack propagation even below a critical stress intensity factor threshold, referred to as subcritical crack growth (Atkinson, 1984). These processes depend on the interaction of several parameters, e.g. in-situ stresses, bedrock petrography, fracture geometries and pore water characteristics. Being significantly favoured by high pore water pressures, lower bound of subcritical fracture propagation velocities vary between a few centimetres and several decimetres per 1000 years (Atkinson, 1984; Atkinson and Meredith, 1987). The application of this fracture mechanical model on unstable slopes would mean that over a longer time period the fracture density and persistence continuously increase. In the long term, this leads to a continuous degradation of the slope stability and to a failure event when the rock mass strength threshold is exceeded. This and a fracture mechanical model for the time-dependent degradation of rock joint cohesion (Kemeny, 2003) imply that after progressive rock strength reduction slope failures may occur very rapidly.

How far the hydrogeological setting, e. g. potential groundwater inflows from the adjacent Fundus valley into the fractured bedrocks in the Köfels area (cf. Fig. 9), contributed to the slope instabilities has not been investigated yet. But the geological field surveys yielded evidence for a close relation between different lithologies, structural setting and slope deformation behaviour. Anisotropic, i.e. structurally controlled, strong rocks, especially at steep slopes featuring high relief energy, generally tend to rapid failure mechanisms such as rock collapses and catastrophic rock slides (Hungr and Evans, 2004). According to this, rapid sliding i.e. the prominent Köfels rockslide developed in areas composed of strong but fractured orthogneisses with dominant sliding planes orientated subparallel to the slope. In contrast, slopes made up by incompetent low-strength paragneisses were not really affected by this rapid event but rather show morphological characteristics of slowly creeping deformations.

These findings were backed up by InSAR analyses, according to which some scarp areas are characterised by low slope deformation rates of around 1 cm per year. Using an extended stack of SAR images the reliability of the InSAR based motion fields can be improved and additionally the annual variation of the motion magnitude in response to seasonal meteorological conditions could be further investigated. Comparable landslide motion fields derived from InSAR analyses were already determined for a few other Alpine slopes in different geological setting (Rott et al., 1999; Rott and Nagler, 2006). Based on this and comparisons with other creeping landslides in paragneissic rock types (e.g. Leobacher and Liegler, 1998; Chwatal et al., 2005; Weißflog, 2007; Zangerl et al., 2007), most likely also the paragneissic Wurzberg started creeping after the retreat of the Late-Pleistocene glaciers and deformed relatively slowly, i.e. without accelerating to a final catastrophic slope collapse.

In summary the Köfels case study demonstrates the capability of geological field surveys and according remote sensing methods, i.e. laserscanner images and spaceborne radar interferometry, to gain information for the assessment of slope instabilities in Alpine environments.

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References

Abele, G., 1974. Bergstürze in den Alpen, ihre Verbreitung, Morphologie und Folgeerscheinung. Wissenschaftliche Alpenvereinshefte, 25, 1-230.

Ampferer, O., 1939. Über die geologischen Deutungen und Bausondierungen des Maurach Riegels im Ötztal. Geologie und Bauwesen, 2, 25-43.

Atkinson, B.K., 1984. Subcritical crack growth in geological materials. Journal of Geophysical Research, 89(B6), 4077-4114.

Atkinson, B.K. and Meredith, P.G., 1987. The theory of subcritical crack growth with applications to minerals and rocks. In: Atkinson, B.K. (ed.), Fracture mechanics of rock, Academic Press, 111–166.

Brandner, R., 1980. Geologische und Tektonische Übersichtskarte von Tirol. Tirol-Atlas, Universitäts-Verlag Wagner, C1, C3.

Brückl, E., 2001. Cause-Effect Models of Large Landslides. Natural Hazards, 23, 291-314.

Brückl, E., Brückl, J. and Heuberger, H., 2001. Present structure and prefailure topography of the giant rockslide of Köfels. Zeitschrift für Gletscherkunde und Glazialgeologie, 37, 49-79.

Chwatal, W., Kirschner, H., Brückl, E. and Zangerl, C., 2005. Geology and 3D seismic structure of the Niedergallmigg-Matekopf mass-movement, Tyrol. Geophysical Research Abstracts, 7, 02566.

Drimmel, J., 1980. Rezente Seismizität und Seismotektonik des Ostalpenraumes. In: Oberhauser, R. (ed.), Der geologische Aufbau Österreichs, Springer, 507-527.

Eberhardt, E., Stead, D., and Coggan, J.S., 2004. Numerical analysis of initiation and progressive failure in natural rock slopes - the 1991 Randa rockslide. International Journal of Rock Mechanics and Mining Sciences, 41, 69–87.

Einstein, H.H., Veneziano, D., Baecher, G.B., and O'Reilly, K.J., 1983. The effect of discontinuity persistence on rock slope stability. International Journal of Rock Mechanics and Mining Sciences, 20/5, 227-236.

Einstein, H.H., 1993. Modern developments in discontinuity analysis - the persistence-connectivity problem. In: Hudson, J.A. (ed.), Comprehensive Rock Engineering, Vol. 3, Pergamon, 193-213.

Einstein, H.H. and Stephansson, O., 2000. Fracture systems, fracture propagation and coalescence. Issue Paper, Proceedings GeoEng 2000, Melbourne.

Eisbacher, G.H. and Brandner, R., 1995. Role of high-angle faults during heteroaxial contraction, Inntal Thrust Sheet, Northern Calcareous Alps, Western Austria. Geologisch-Paläontologische Mitteilungen Innsbruck, 20, 389-406.

Erismann, T.H., Heuberger, H., and Preuss, E. , 1977. Der Bimsstein von Köfels (Tirol), ein Bergsturz-"Friktionit". Tschermaks Mineralogisch-Petrographische Mitteilungen, 24, 67-119.

Erismann, T.H. and Abele, G., 2001. Dynamics of rockslides and rockfalls. Springer, 316 pp.

Geitner, C. 1999. Sedimentologische und vegetationsgeschichtliche Untersuchungen an fluvialen Sedimenten in den Hochlagen des Horlachtales (Stubaier Alpen/Tirol) – Ein Beitrag zur zeitlichen Differenzierung der fluvialen Dynamik im Holozän. Münchener Geographische Abhandlungen, Reihe B, 32, 1-247.

Goodmann, R.E. and Kieffer S.D., 2000. Behavior of Rock in Slopes. Journal of Geotechnical and Geoenvironmental Engineering, 126/8, 675-684.

Hammer, W., 1929. Geologische Spezialkarte der Republik Österreich 1:75.000, 5146 Ötzthal. Verlag Geologische Bundesanstalt, Wien.

Heidbach, O., Tingay, M., Barth, A., Reinecker, J., Kurfeß, D., and Müller, B., 2008. The World Stress Map database release 2008, doi:10.1594/GFZ.WSM.Rel2008, online at: www.world-stress-map.org.

Hermanns, R.L., Blikra, L.H., Naumann, M., Nilsen, B., Panthi, K.K., Stromeyer, D., and Longva, O., 2006. Examples of multiple rock-slope collapses from Köfels (Ötz valley, Austria) and western Norway. Engineering Geology, 83, 94-108.

Heuberger, H., 1966. Gletschergeschichtliche Untersuchungen in den Zentralalpen zwischen Sellrain- und Ötztal. Wissenschaftliche Alpenvereinshefte, 20, 1-126.

Heuberger, H., 1975. Das Ötztal. Bergstürze und alte Gletscherstände, kulturgeographische Gliederung. Innsbrucker Geographische Studien, 2, 213-249.

Heuberger, H., 1994. The giant landslide of Köfels, Ötztal, Tyrol. Mountain Research and Development, 14, 290-294.

Hoinkes, G., Purtscheller, F, and Tessadri, R., 1982. Polymetamorphose im Ostalpin westlich der Tauern (Ötztaler Masse, Schneeberger Zug, Brennermesozoikum). Zusammenfassung der petrographischen Neuergebnisse. Geologisch-Paläontologische Mitteilungen Innsbruck, 12, 95-113.

Hoinkes, G. and Thöni, M., 1993. Evolution of the Ötztal-Stubai, Scarl-Campo and Ulten Basement Units. In: Raumer, J.F. and Neubauer, F. (eds.), The Pre-Mesozoic Geology in the Alps, Springer, pp. 485-494.

Hungr, O. and Evans, S.G., 2004. The occurrence and classification of massive rock slope failure. Felsbau, 22, 16-23.

Ivy-Ochs, S., Heuberger, H., Kubik, P.W., Kerschner, H., Bonani, G., Frank, M., and Schlüchter, C., 1998. The age of the Köfels event. Relative, 14C and cosmogenic isotope dating of an early Holocene landslide in the Central Alps (Tyrol, Austria). Zeitschrift für Gletscherkunde und Glazialgeologie, 34, 57–68.

Kemeny, J., 2003. The Time-Dependent Reduction of Sliding Cohesion due to Rock Bridges Along Discontinuities: A Fracture Mechanics Approach. Rock Mechanics and Rock Engineering, 36/1, 27-38.

Kieffer, S.D., 2003. Rotational Instability of Hard Rock Slopes. Felsbau, 21/2, 31-38.

Klebelsberg, R., 1951. Das Becken von Längenfeld im Ötztal. Schlern-Schriften, 77, 399-422.

Kubik, P.W., Ivy-Ochs, S., Masarik, J., Frank, M., and Schlüchter, C., 1998. ^{10}Be and ^{26}Al production rates deduced from an instantaneous event within the dendro-calibration curve, the landslides of Köfels, Ötz valley, Austria. Earth and Planetary Science Letters, 161, 231-241.

Kubik, P.W., Ivy-Ochs, S., and Kerschner, H., 2007. Multiple prehistoric landslides at Köfels (Austria): Timing by cosmogenic ^{10}Be. Geophysical Research Abstracts, 9, 02177, EGU2007.

Kurat, G. and Richter, W., 1972. Impaktite von Köfels. Tschermaks Mineralogisch-Petrographische Mitteilungen, 17, 23-45.

Leobacher, A. and Liegler, K. 1998. Langzeitkontrolle von Massenbewegungen der Stauraumhänge des Speichers Durlaßboden. Felsbau, 16/3, 184-193.

Masch, L., Wenk, H.R., and Preuss, E., 1985. Electron microscopy study of hyalomylonites - evidence for frictional melting in landslides. Tectonophysics, 115, 131-160.

Ortner, H., 2003. Local and far field stress-analysis of brittle deformation in the western part of the Northern Calcareous Alps. Geologisch-Paläontologische Mitteilungen Innsbruck, 26, 109-136.

Pagliarini, L., 2008. Strukturelle Neubearbeitung des Tschirgant und Analyse der lithologisch-strukturell induzierten Massenbewegung (Tschirgant Bergsturz, Nördliche Kalkalpen, Tirol). Diploma Thesis, Univ. Innsbruck, 90 pp.

Pichler, A., 1863. Zur Geognosie Tirols. II. Die vulkanischen Reste von Köfels. Jahrbuch der Kaiserlich Königlichen Reichsanstalt, 13, 591-594.

Poscher, G. and Patzelt, G., 2000. Sink-hole Collapses in Soft Rocks. Felsbau, 18, 36-40.

Prager, C., Zangerl, C., Patzelt, G., and Brandner, R, 2008. Age distribution of fossil landslides in the Tyrol (Austria) and its surrounding areas. Natural Hazards and Earth System Sciences, 8/2, 377-407.

Prager, C., Ivy-Ochs, S., Ostermann, M., Synal, H.-A., and Patzelt, G., 2009. Geology and radiometric 14C-, 36Cl- and Th-/U-dating of the Fernpass rockslide (Tyrol, Austria). Geomorphology, 103, 93-103.

Preuss, E.,1974. Der Bimsstein von Köfels/Tirol: die Reibungsschmelze eines Bergsturzes. Jahrbuch des Vereins zum Schutze der Alpenpflanzen und -Tiere, 39, 85-95.

Preuss, E.,1986. Gleitflächen und neue Friktionitfunde im Bergsturz von Köfels im Ötztal, Tirol. Material und Technik, 3, 169-174.

Purtscheller, F., 1978. Ötztaler und Stubaier Alpen. Sammlung Geologischer Führer, 53, 1-128, Bornträger.

Purtscheller, F. and Rammlmair, D., 1982. Alpine Metamorphism of Diabase Dikes in the Ötztal-Stubai Metamorphic Complex. Tschermaks Mineralogisch-Petrographische Mitteilungen, 29, 205-221.

Purtscheller, F., Pirchl, T., Sieder, G., Stingl, V., Tessadri, T., Brunner, P., Ennemoser, O., and Schneider, P., 1995. Radon emanation from giant landslides of Koefels (Tyrol, Austria) and Lang Tang Himal (Nepal). Environmental Geology, 26, 32-38.

Purtscheller, F., Stingl, V., Tessadri, R., Mair, V., Pirchl, T., Sieder, G., Brunner, P., Schneider, P., and Ennemoser, O., 1997. Geologisch-petrographische Ursachen der Radonanomalie in Umhausen (Tirol). Mitteilungen der Österreichischen Geologischen Gesellschaft, 88, 7-13.

Rott, H., Scheuchl, B., Siegel, A., and Grasemann, B., 1999. Monitoring very slow slope movements by means of SAR interferometry: a case study from a mass waste above a reservoir in the Ötztal Alps, Austria. Geophysical Research Letters, 26, 1629-1632.

Rott, H. and Nagler, T., 2006. The contribution of radar interferometry to the assessment of landslide hazards. Advances in Space Research, 37, 710-719.

Sassi, F.P., Cesare, B., Mazzoli, C., Peruzzo, L., Sassi, R., and Spiess, R., 2004. The crystalline basements of the Italian Eastern Alps: a review of the metamorphic features. Periodico di Mineralogia, 73, 23-42.

Schmid, S.M., Fügenschuh, B., Kissling, E., and Schuster, R., 2004. Tectonic map and overall architecture of the Alpine orogen. Eclogae Geologicae Helvetiae, 97, 93-117.

Senarclens-Grancy, W., 1958. Zur Glazialgeologie des Oetztales und seiner Umgebung. Mitteilungen der Geologischen Gesellschaft Wien, 49, 257-314.

Sieder, G. and Pirchl, T., 1994. Geologische und petrographische Untersuchungen im Raum Umhausen/Ötztal mit besonderer Berücksichtigung der Radonproblematik. Diploma Thesis, Univ. Innsbruck, 91 pp.

Sørensen, S.-A. and Bauer, B., 2003. On the dynamics of the Köfels Sturzstrom. Geomorphology, 54, 11-19.

Suess, F.E., 1937. Der Meteor-Krater von Köfels bei Umhausen im Ötztale, Tirol. Neues Jahrbuch für Mineralogie, Geologie und Paläontologie, Abh., 72, 98-155.

Tentschert, E., 1998. Das Langzeitverhalten der Sackungshänge im Speicher Gepatsch (Tirol, Österreich). Felsbau, 3, 194-200.

Tiris, 2009. Laserscanning Tirol. http://tiris.tirol.gv.at/web/index.cfm

Van Gool, J.A.M., Kemme, M.M.J., and Schreurs, G.M.M.F., 1987. Structural investigations along an E-W cross-section in the southern Ötztal Alps. In: Flügel H.W. and Faupl P. (eds.), Geodynamics of the Eastern Alps, Deuticke, pp. 214-222.

Weidinger, J.T. and Korup, O., 2009. Frictionite as evidence for a large Late Quaternary rockslide near Kanchenjunga, Sikkim Himalayas, India - implications for extreme events in mountain relief destruction. Geomorphology, 103, 57–65.

Weißflog, M., 2007. Ursachen und Phänomene des Talzuschubs Stupfarri-Kaltenbrunn/Vorderes Kaunertal (Tirol/Österreich). Geomorphologische Untersuchung einer großräumigen Massenbewegung im Ötztal-Stubai-Kristallin. Diploma Thesis, TU München, 109 pp.

Zangerl, C., Eberhardt, E., Schönlaub, H., and Anegg, J., 2007. Deformation behaviour of deep-seated rockslides in crystalline rock. Proceedings of the 1st Canada – U.S. Rock Mechanics Symposium, Vancouver, pp. 901-908.

Zhang, Z.-Y., Chen, S.-M., and Tao, L.-J, 2002. 1983 Sale Mountain landslide, Gansu Province, China. In: Evans S. G. and DeGraff J. (eds.), Catastrophic Landslides: Effects, Occurrence, and Mechanisms, Reviews in Engineering Geology, 15, Geol. Soc. Am., pp. 149-163.

Zötl, J. and Goldbrunner, J.E., 1993. Die Mineral- und Heilwässer Österreichs, Geologische Grundlagen und Spurenelemente. Springer, 329 pp.

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